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Meltwater flow through a rapidly deglaciating glacier and foreland catchment system: Virkisjökull, SE Iceland

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ABSTRACT

Virkisjökull is a rapidly retreating glacier in south-east Iceland. A proglacial lake has formed in the last ten years underlain by buried ice. In this study we estimate water velocities through the glacier, proglacial foreland and proglacial river using tracer tests and continuous meltwater flow measurements. Tracer testing from a glacial moulin to the glacier outlet in September 2013 demonstrated a rapid velocity of 0.58 m s^{-1}. This was comparable to the velocity within the proglacial river, also estimated from tracer testing. A subsequent tracer test from the same glacial moulin under low flow conditions in May 2014 demonstrated a slower velocity of 0.07 m s^{-1}. The glacier outlet river sinks back into the buried ice, and a tracer test from this sink point through the proglacial foreland to the meltwater river beyond the lake indicated a velocity of 0.03 m s^{-1}, suggesting that an ice conduit system within the buried ice is transferring water rapidly beneath the lake. Ground penetrating radar profiles confirm the presence of this buried conduit system. This study provides an example of rapid deglaciation being associated with extensive conduit systems that enable rapid meltwater transfer from glaciers through the proglacial area to meltwater rivers.

Key words | buried ice, dye tracing, Iceland, proglacial area, Virkisjökull

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INTRODUCTION

Glaciers in Iceland currently cover ~11,000 km^2 (Jóhannesson *et al.* 2013), but climate change is causing widespread glacial retreat (Haugen & Iversen 2005; Halldórsdóttir *et al.* 2006; Jóhannesson *et al.* 2006; Fenger 2007; Aðalgeirsdóttir *et al.* 2011), particularly since 1995 (Björnsson *et al.* 2013). Deglaciation results in the rapid evolution of englacial conduits (Nienow *et al.* 1998) and increased development of crevasses and moulins (Catania & Neumann 2010). The IPCC (2014) reports that increased glacier melt is resulting in changes to proglacial river systems with consequences for the management of water resources and hazards. In some catchments, an overall increase in river discharge has been observed due to the increase in meltwater volume (Björnsson & Pálsson 2008; Nolin *et al.* 2010). In particular, there is likely to be an increase in winter flows due to increased winter temperatures (Fountain & Tangborn 1985), and predictions have been made of a transition from ephemeral to perennial river flows (Jóhannesson *et al.* 2007). Deglaciation also causes other changes in proglacial forelands. Buried ice may be left behind as glaciers retreat (French & Harry 1990; Evans & England 1992; Everest & Bradwell 2003). Proglacial lakes may also be rapidly formed and disappear (Kirkbride 1993; Ageta & Iwata 2000; Bennett & Evans 2012).

Virkisjökull is a rapidly deglaciating maritime glacier in south-east Iceland. Sequential field photographs and annual moraines show that there has been substantial retreat over the period 2008–2013 averaging 35 m a^{-1} (Bradwell *et al.*

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2013). The glacier margin has retreated nearly 500 m since 1996, and there has been a decrease in the glacier surface elevation of 8 m a^{-1} in the lowest reaches since 2012. The rate of retreat is accelerating, with an increase from 14 m a^{-1} of retreat between 1990 and 2004 to 33 m a^{-1} of retreat between 2005 and 2011 (Bradwell *et al.* 2013). The proglacial foreland is changing in response to rapid deglaciation, and is characterised by extensive areas of buried ice and a growing proglacial meltwater lake, formed within the last ten years.

The aim of this study was to use tracer tests, river discharge measurements, and ground penetrating radar (GPR) to characterise the glacial and proglacial hydrology of this rapidly deglaciating system, and to determine the water velocity through the glacier and proglacial area.

Site description and overview of testing

Virkisjökull comprises two glacier arms that are split by a rocky ridge, and join at the terminus (Figure 1). The British Geological Survey, in collaboration with the Icelandic Meteorological Office, have been working in the Virkisjökull catchment since 2009: monitoring the rapid retreat of the glacier (Bradwell *et al.* 2013); investigating mechanisms for this rapid retreat (Phillips *et al.* 2013, 2014); monitoring the hydrology of the glacier (MacDonald *et al.* 2016); and measuring mass balance (Flett 2016). The glacier is drained by the river Virkisá which has a catchment area of approximately 31 km^2 and extends southwestwards from the summit crater of Öræfajökull. There is a discrete river discharging from the terminus on the east side of the glacier (Point 3 on Figure 1), but on the west side there is no obvious channelised discharge.

Tracer tests were undertaken from glacial moulins on the east and west arms of the glacier (Points 1 and 2 on Figure 1) to determine whether they are connected to the outlet at the glacier terminus (Point 3 on Figure 1), and investigate meltwater velocities. GPR was used to investigate the location of the conduits in the lower ablation zone and proglacial area. These data were set within the context of the continuous measurements of river discharge (MacDonald *et al.* 2016).

Beyond the terminus, the proglacial foreland is characterised by an area of buried ice. Daily photographs show that during typical flows, the river discharging from the terminus (Point 3 on Figure 1) flows intermittently across the proglacial foreland surface for approximately 50 m before it sinks into the buried ice via a collapse feature. This river does not resurface but collapse features in the sediment overlying the ice are prevalent. To the west the buried ice is overlain by a lake (Figure 1). The lake area is complex, with debris piles forming islands. Unstable ground prevents access to some areas of the lake shore. On the far side of the lake, an outlet feeds the proglacial river. At the lake outlet, the proglacial area is geologically constrained by low permeability bedrock, therefore, the majority of melt water discharges through this area. A tracer test was undertaken to measure the transit time through the proglacial area. GPR surveys were undertaken prior to this study in the proglacial area to investigate the collapse of the glacier margin (Phillips *et al.* 2013). They are used here to determine whether conduits are present within the buried ice.

FIELD METHODS

Proglacial river discharge measurements

River discharge was measured over three years at an automatic gauging station, 2.92 km downstream of the lake outlet (MacDonald *et al.* 2016; Point 6 on Figure 1). This was to determine temporal changes in meltwater discharge, and establish whether melting occurs on a perennial basis. These data are also used to show river conditions during the tracer tests.

Water level was monitored continuously using submersible level transmitters attached to the adjacent road bridge. An Ott Kalesto V surface velocity sensor was also deployed to indicate when changes in channel morphology were likely to have affected the stage–discharge rating. Velocities were measured with an Ott C-31 current meter by wading and with a 3 m pole from the bridge during high flows. The resultant 56 flow gaugings between 2011 and 2014 allowed the stage data to be converted to flow at 15 minute intervals for the period 2011-09-16 until 2014-11-15. As ice affects the stage–discharge relationship, photographs were taken three times daily using an automated system, at

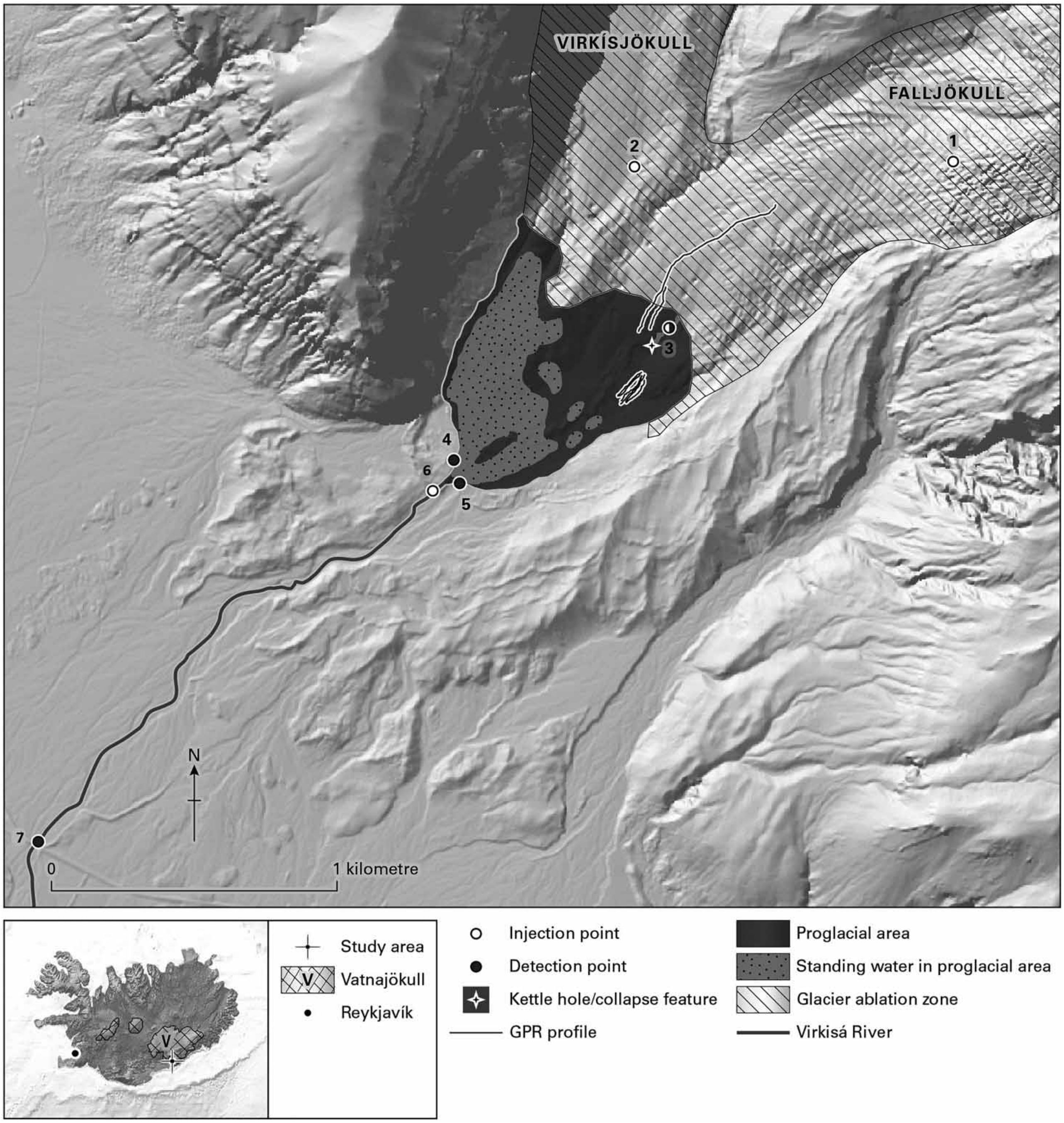

Figure 1 | The lower part of the Virkisjökull catchment including the lower ablation zone and the proglacial lake area and outlet river based on 2013 extent. Points labelled on the map are: (1) east arm injection moulin; (2) west arm injection moulin May 2014 and August 2014; (3) glacier snout outflow monitoring point/proglacial river sink injection point; (4) lake outlet west monitoring point; (5) lake outlet east monitoring point; (6) proglacial river monitoring point (for proglacial foreland tracer test); (7) river dye injection point.

9:00, 12:00 and 15:00, to identify ice development in the channel. Periods when ice was present in the channel or around the banks were removed from the discharge record.

Tracer tests

In all tracer tests a 40% sodium fluorescein or rhodamine WT dye solution was used. Sodium fluorescein has a high photochemical decay rate under natural light (Smart & Laidlaw 1977), and therefore was only used for tests that had minimal exposure to light.

Tracer tests were carried out in September 2013 and May 2014 from a moulin on the east arm of Virkisjökull (Point 1 on Figure 1), 1.5 km from the terminus (Figure 2). A single tracer test was carried out from two large moulins on the western arm of the glacier in May and August 2014 (Point 2 on Figure 1). These moulins were 840 m and 670 m from the glacial outlet on the east side of the terminus (Point 3 on Figure 1). Monitoring for all these tracer tests was carried out in the river just downstream from this discharge point (Point 3 on Figure 1).

In September 2014, tracer was injected into the river that emerges and then sinks at the glacier terminus (Point 3 on Figure 1). The east and west side of the lake outlet channel were monitored (Points 4 and 5 on Figure 1).

A final tracer test was conducted to determine the velocity in the proglacial river. Tracer was injected downstream from the lake outlet (Point 6 on Figure 1), with monitoring 2.92 km downstream (Point 7 on Figure 1).

Figure 2 | The eastern arm glacier injection moulin in September 2013 (a) and April 2014 (b).

GGUN-FL *in situ* field fluorometers designed at Neuchatel University were used at all monitoring sites to measure fluorescence and turbidity at 2 minute intervals. The fluorometers were calibrated to standard concentrations before field work commenced. Monitoring was started 24 hours before the tests and continued for a minimum of 2 days except for occasions with high flows.

Turbidity increases fluorescence, and fluctuates in response to discharge (Wilson *et al.* 1986). Background fluorescence is therefore highly variable in glacial environments where discharge varies considerably, due to both diurnal fluctuations in melting, and in response to rainfall (Schnegg 2002). Turbidity measurements were collected concurrently with fluorescence data, and used to correct the data for variations in background fluorescence. For each breakthrough curve the relationship between turbidity (ppb) and fluorescence (ppb) was established during the background period prior to tracer injection using XY scatter plots. Correlations were good, and r^2 ranged from 0.70 to 0.99. The equation defining the relationship was used to determine a corrected background fluorescence value (ppb) which was then subtracted from the measured fluorescence to provide a corrected breakthrough curve. Tracer recoveries were not calculated because it was not safe to measure the river discharge at the detection points.

Dispersivity is a measure of the amount by which dye becomes dispersed (m) as it moves downstream, and was calculated using the equation in Seaberg *et al.* (1988; Equation (4), p. 222).

GPR

GPR surveys were performed at Virkisjökull in April 2012 and 2013 as part of an ongoing study into the structural glaciology (Phillips *et al.* 2013, 2014). The results of these surveys are used here to infer water flow paths within the glacier. A GPR survey was also conducted in the proglacial area in September 2012. A PulseEKKO Pro system with 50 MHz and 100 MHz antennae was used. Antennae were aligned perpendicular to travel direction and towed manually across the surface, with the radar being triggered every 0.25 m by an odometer wheel. Where the ice surface was fractured, the antennae were moved stepwise and the radar was triggered manually. Positional data were stored alongside GPR trace data using a standalone Novatel SMART-V1 GPS antenna. Raw GPR data were processed in EKKO View Deluxe. The processing consisted of applying a dewow filter, 2-D migration (for clean ice surveys), SEC (spreading and exponential compensation) gain, and topographic correction. For the clean glacier ice, a radar wave velocity of 0.156 m ns^{-1}, previously calculated for Virkisjökull, was used (Murray *et al.* 2000).

RESULTS

River discharge

Measured river discharge fluctuations at Virkisá are typical of sub-Arctic meltwater rivers which are dominated by seasonal and diurnal temperature fluctuations (Shaw *et al.* 2011). The mean summer discharge at Virkisá is 5.3–7.9 $m^3 s^{-1}$, and the winter mean is 1.6–2.4 $m^3 s^{-1}$. The highest flow recorded during the period of measurement was 72 $m^3 s^{-1}$ in October 2014, and the lowest recorded flow was 0.3 $m^3 s^{-1}$ (MacDonald *et al.* 2016). There is a strong seasonal variation in discharge which increases between May and September and decreases between September and November (Figure 3). Photographs taken three times per day at the road-bridge and direct measurements indicate that the river flows throughout the year, despite ice developing around the banks and daytime temperatures in winter falling below 0 °C for 5 consecutive days or more. More detailed analysis of the hydrographs are given in MacDonald *et al.* (2016). Further details on the river discharge conditions during each test are given in their specific sections below.

Tracer tests

Table 1 summarises the main findings for direct comparison between the systems. Following the east arm glacier tracer test in September 2013, tracer breakthrough at the glacier terminus (1.5 km from the injection point) was rapid, occurring 50 minutes after dye injection (Figure 4(a)). Peak concentrations were approximately 58 minutes after injection. Tracer concentrations declined to below background levels 4 hours and 33 minutes after the dye injection. Monitoring stopped 11 hours and 43 minutes after injection when the water level dropped below the level of the fluorometer because the water was diverted naturally into a different channel. This effect did not occur during any other tests. This test indicates that near the end of the main ice ablation season (defined as the day with the first extensive snow fall) the meltwater transmission to the glacier margin (based on the time to peak of the tracer test) was 0.58 $m s^{-1}$. The test was conducted during moderately high discharge measured at the proglacial river (5.5 $m^3 s^{-1}$). Dispersivity was 4.7 m.

An identical tracer test at the start of the ablation season was undertaken in May 2014, when the discharge in the proglacial river was approximately 2 $m^3 s^{-1}$. At this time, flow into the same moulin on the eastern arm of Virkisjökull was also substantially lower. Tracer breakthrough occurred 5 hours and 18 minutes after the dye injection (Figure 4(a)). Peak tracer concentrations occurred 5 hours and 36 minutes after injection. Tracer concentrations had not returned to background when monitoring stopped 20 hours after injection due to high flows which put the fluorometer at risk. The meltwater velocity (based on the time to peak of the tracer test) was 0.07 $m s^{-1}$ and the dispersivity was 35.9 m, which was higher than in the previous test.

The western arm tracer test in May 2014 resulted in no observable breakthrough curve at the glacier terminus (Figure 4(b)). Monitoring stopped 11 hours and 57 minutes after injection because of movement of the outlet channel. While it is possible that the tracer breakthrough could have occurred after monitoring stopped, this is unlikely given that the monitoring continues for substantially longer than the time taken for tracer breakthrough from the moulin on the eastern arm. The east and west glacier arm tests were carried out during the same period in May 2014, and under similar discharge conditions, when there were flows of approximately 2 $m^3 s^{-1}$ in the proglacial river. It is likely that there was sufficient flow in the western injection moulin to flush the tracer through the system. In addition, the moulin on the western arm is much closer to the glacier terminus monitoring point than the moulin on the eastern arm.

In August 2014, a repeat tracer test was carried out from the two moulins on the western arm of the glacier but there was also no observable breakthrough curve (Figure 4(c)), despite 2 days of monitoring following

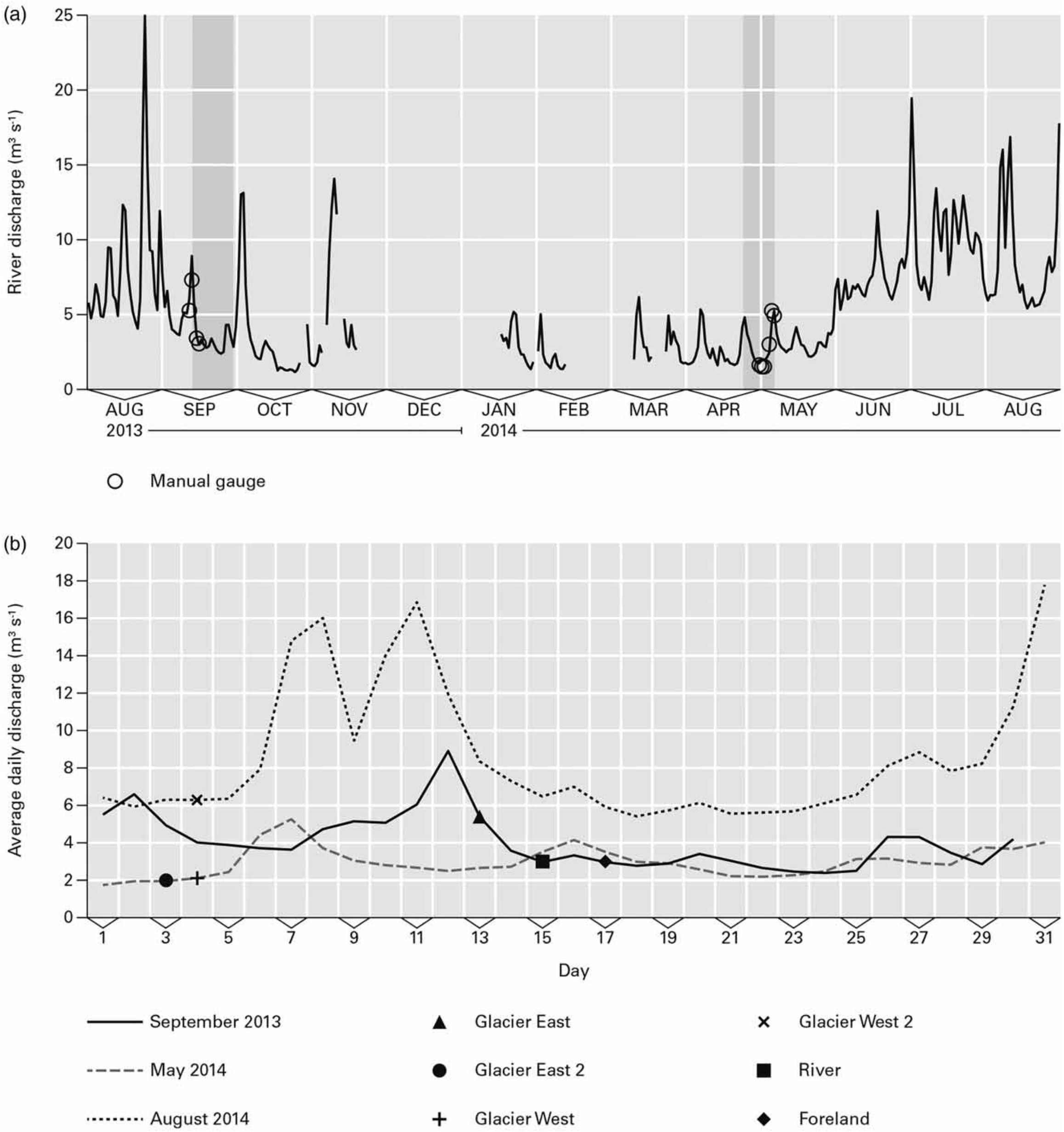

Figure 3 | (a) River discharge record from August 2013 to August 2014 showing areas in grey when tracer tests were carried out; circles indicate times of manual river gaugings. Gaps in the record are dates removed due to channel ice. (b) Detail of the river discharge measurements during the tracer tests.

injection and high flow rates of approximately 6 $m^3 s^{-1}$ in the proglacial river during the test (Figure 3). This suggests that the drainage from the western arm of the glacier is not connected to the outlet at the glacier terminus but flows directly into the proglacial area through the buried ice and lake.

During the proglacial area tracer test, tracer was injected into the sinking river at the glacier terminus (Point 3 on Figure 1). A breakthrough curve was obtained at the lake outlet west bank monitoring site at the south end of the proglacial foreland area (Point 4 on Figure 1). Tracer breakthrough was at 19:00 on 2013-09-27, 7.5 hours after injection (Figure 5(a)). The peak concentration occurred at 20:30, 9 hours after injection (Table 1). There was no tracer breakthrough at the lake outlet east monitoring point (Point 5 on Figure 1) (Figure 5(b)).

Table 1 | Summary of all tracer tests performed at Virkisjökull and their main findings

Location and test name	Injection time and date	Dye	Dye (g)	Distance (m)	Time to peak dye conc. (min)	Velocity (m s^{-1})	Dispersivity (m)	Post-injection monitoring (hr)
Glacier east	12:04, 2013-09-12	Fluorescein	500	1,500	58	0.57	4.7	11.5
Glacier east 2	14:16, 2014-05-03	Fluorescein	500	1,500	332	0.07	35.9	12.5
Glacier west	16:02, 2014-05-04	Rhodamine WT	400	668	NA	NA	NA	12
Glacier west 2	11:18, 2014-08-04	Fluorescein	580	839	NA	NA	NA	48
Foreland	16:48, 2013-09-17	Rhodamine WT	2,199	1,000	450	0.03	29.4	72
River	11:45, 2013-09-15	Rhodamine WT	128	2,920	90–115	0.6	5–7.3	4

Velocity is calculated using the distance and time to peak.

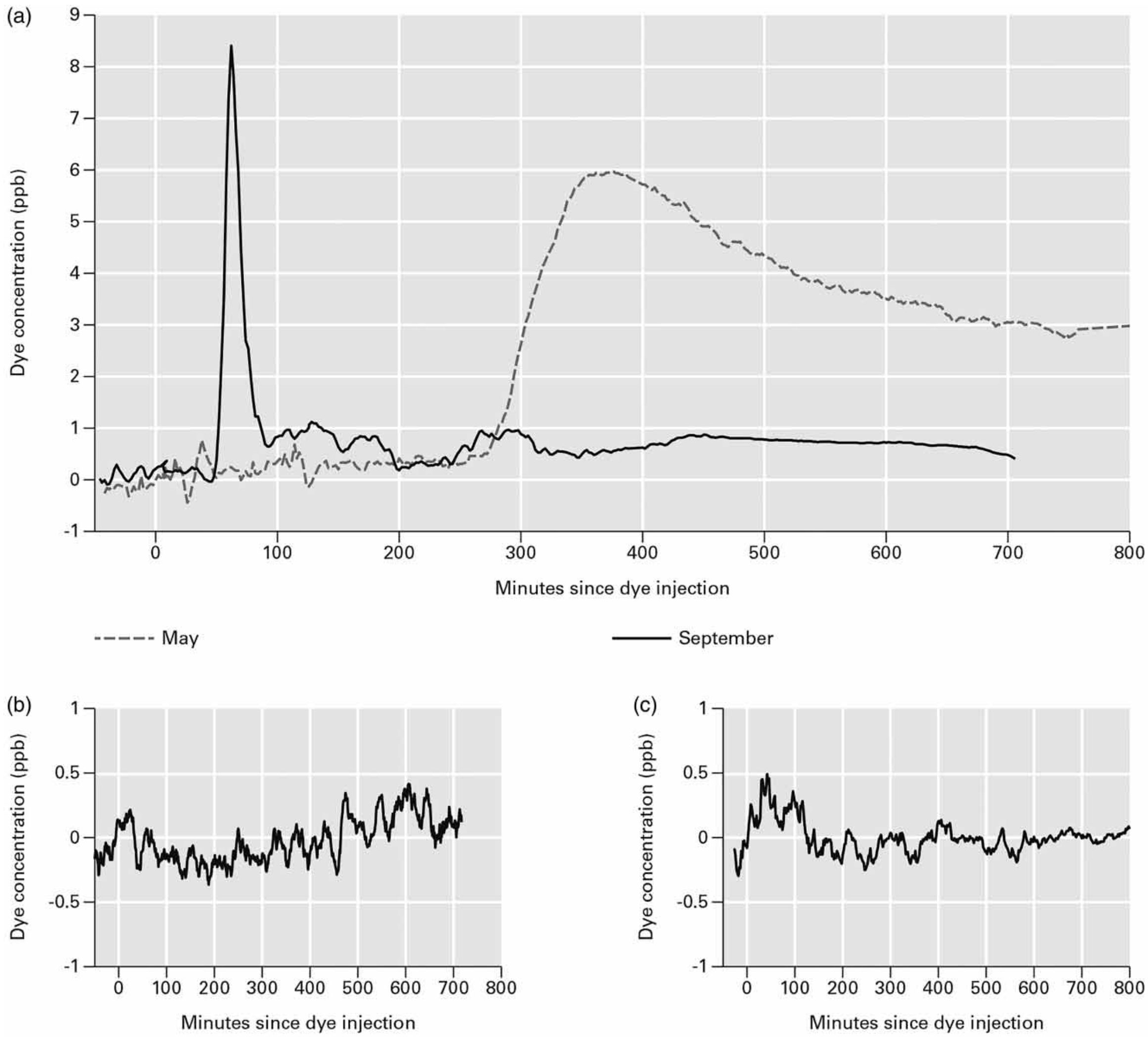

Figure 4 | (a) Tracer breakthrough curves during the glacier tracer tests on the eastern glacier arm in September 2013 and May 2014. (b) Results from the test undertaken from the moulin on the western glacier arm in May 2014 showing no breakthrough. (c) Results from the test undertaken from the moulin on the western glacier arm in August 2014 showing no breakthrough. All figures are corrected for background fluorescence and turbidity.

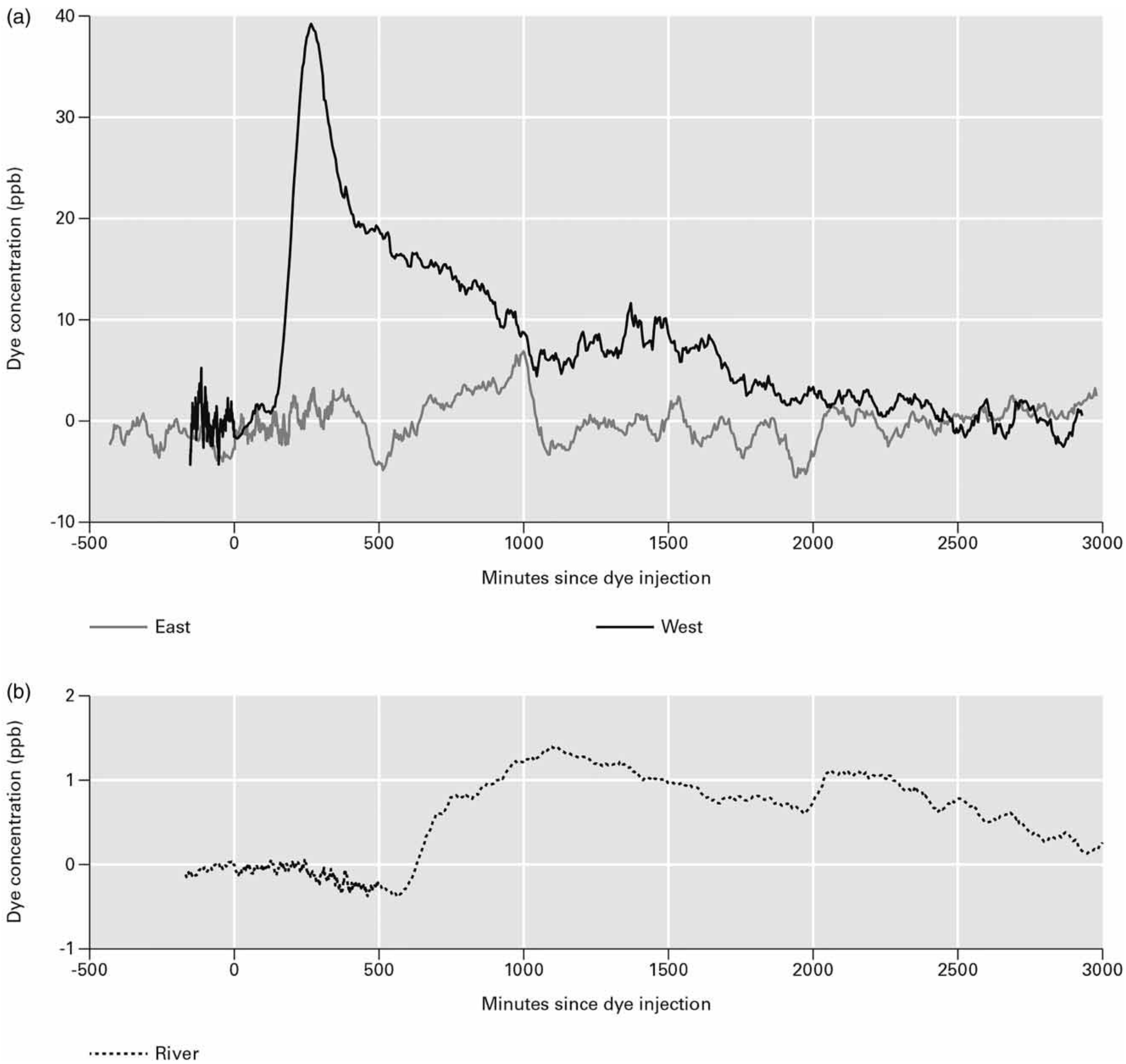

Figure 5 | Rhodamine WT breakthrough curves during the proglacial tracer test: (a) at the lake outlet channel west, at the lake outlet channel east; and (b) at the downstream river gauging station. All figures are corrected for background fluorescence and turbidity.

The tracer test in the proglacial river in September 2013 (injected at Point 7 and detected at Point 6 on Figure 1) indicated a velocity of 0.6 m s^{-1}. The flow was lower and decreasing during the river and proglacial foreland tracer tests (4–3 $m^3 s^{-1}$). Dispersivity was 29.4 m.

GPR

Sub-horizontal to gently up-ice dipping reflective surfaces within the glacier are apparent in profiles for the lower parts of the glacier (Figure 6(a)). In several areas their polarity is reversed, indicating a higher dielectric permittivity and lower wave velocity in the material below interface, suggesting the presence of water, or wet sediment. These sub-horizontal reflectors are longitudinally continuous for distances in excess of 100 m; they have been interpreted as thrust planes (Phillips *et al.* 2013), where the fractured ice potentially provides a zone for water flow and conduit development. Field observations (Figure 6(b)) confirmed the presence of wet, graded (waterlain) sediment, and conduits in one of these thrust planes. A prominent down-glacier dipping reflector extends from the glacier surface, where it occurs in association with three moulins, down to the glacier bed approximately 50 m below the

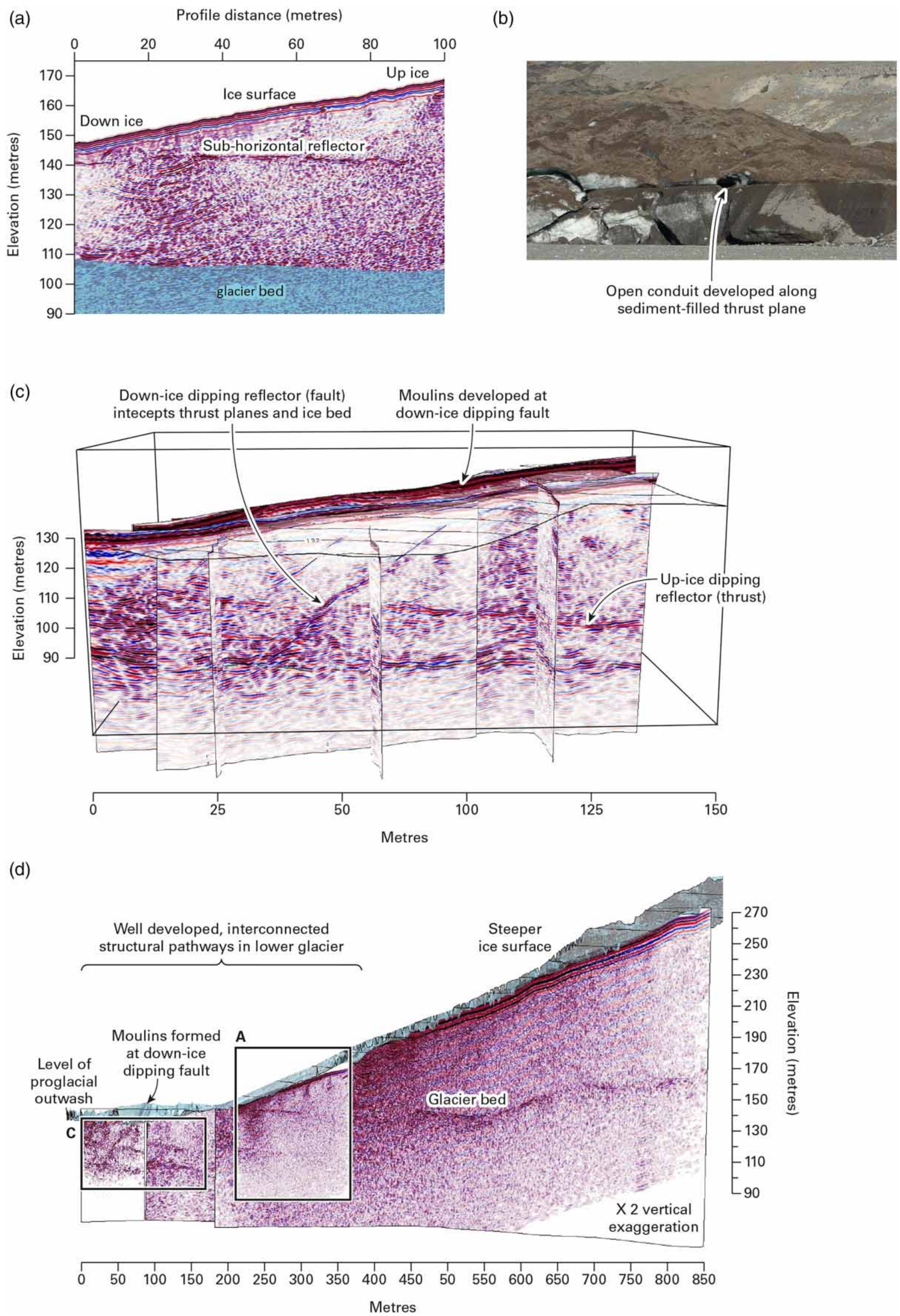

Figure 6 | (a) Continuous sub-horizontal reflector interpreted as thrust plane. (b) Field photograph at debris covered ice margin showing a conduit developed within a sediment-filled thrust plane. (c) Enlarged composite image showing up-glacier and down-glacier dipping reflectors in the lower glacier, interpreted as thrust and fault planes. (d) Composite of glacier GPR surveys showing a zone in the lower glacier with clear reflectors representing a well-developed englacial structural network.

surface (Figure 6(c)). The reflector joins a zone of lateral fractures at the ice surface, and is interpreted as part of a down-glacier dipping fault system where part of the glacier is collapsing (Phillips *et al.* 2013). The location of moulins at the fault zone allows the system to act as an effective water flow route. Collectively, the results from the glacier suggest that there is a pattern of conduit formation in the lower glacier which is associated with ice structures (thrusts, faults), ice-surface topography and the position of moulins which occur predominately on the eastern side of the glacier where there may be a high meltwater input to the fault and thrust plane network (Figure 6(d)).

GPR profiles in the proglacial area (Figure 7(a)) are characterised by an upper unit of horizontal and gently undulating reflectors overlying a generally less reflective unit (Figure 7(b)). Field observations confirm that the upper unit is stratified outwash sand and gravels which are 1–2 m in thickness, and that the lower less reflective unit is buried ice (Figure 7(b) and 7(c)). The top of the buried ice is characterised by a number of hyperbolae, which may represent water-filled conduits or cavities close to the ice surface. The ice depth at the glacier terminus (not including the area of unmoving buried ice in the foreland) was approximately 40 m in 2012. The base of the ice is marked by a transition back to higher amplitude reflectors (Figure 7(b)). Clear, reversed polarity hyperbolae occur in several places in the ice (Figure 7(d)). These are interpreted as water-filled conduits and, in places, are associated with a thickened zone of chaotic reflectors in the sands and gravel above, representing collapsing ground (Figure 7(d)). A marked zone of muted or absent reflections was observed in a number of profiles where they crossed a distinct linear zone that had been particularly affected by collapse holes. The exact reason for poor reflection in this zone is not known, but it may be related to a turbulent subterranean river that was observed sinking underground in this zone at the time of survey (Figure 7(a)). Field observations indicate that kettle holes and collapse features in the proglacial zone intercept a freely draining system as meltwater from the terminus is regularly redirected into one of these features. Collectively, the radar data and observations from the proglacial area demonstrate the presence of an extensive mass of ice buried, with numerous conduits and voids, beneath the outwash sands and gravels.

DISCUSSION

The glacial drainage system

Meltwater velocities through the glacier from the eastern arm are rapid at the end of the main melt season ($0.58\ \mathrm{m\ s^{-1}}$) and are comparable to the upper range of velocities of other glacial tracer tests (Table 2). The velocity is also almost identical to the velocity of $0.6\ \mathrm{m\ s^{-1}}$ measured in the meltwater river channel. The second tracer test from the eastern arm in May 2014, at the beginning of the ice ablation season, demonstrates a lower velocity of $0.07\ \mathrm{m\ s^{-1}}$, but again with the range measured in other glacier tracer tests (Table 2). There is, therefore, an order of magnitude change in velocity between the two tests. It is interesting to compare these tracer results in glaciers to tracer tests in karst conduits and caves, where many different tests are routinely carried out to trace groundwater flow. Worthington & Ford (2009) compiled velocities from 3,015 karst tracer tests and found a median velocity of $0.02\ \mathrm{m\ s^{-1}}$, less than most of the velocities measured in glaciers, even those at the start of the ice ablation season.

Below we discuss four possible reasons for the increase in measured velocity from May to September.

1. There could have been a significant change in the conduit geometry during the months between the tracer tests. The flow would still be in the same conduit, but with different geometry, for example, due to melting, freezing, physical erosion of ice, erosion of debris or deposition of debris.
2. There could be a change in the type of ice conduit system, with a transition from a simple system of well-connected large conduits (termed 'channelised' flow) to a network of small conduits and fissures in the ice (termed 'distributed' flow). Previous studies have suggested that changes in meltwater velocity may be due to a change from a channelised system to a distributed one (Seaberg *et al.* 1988; Willis *et al.* 1990; Fountain 1993; Hock & Hooke 1993; Nienow *et al.* 1998). Generally, meltwater velocities lower than $0.4\ \mathrm{m\ s^{-1}}$ have been interpreted as flow in a distributed system (Nienow *et al.* 1998; Mair *et al.* 2002; Hubbard & Glasser 2005). However, velocities in karst conduit

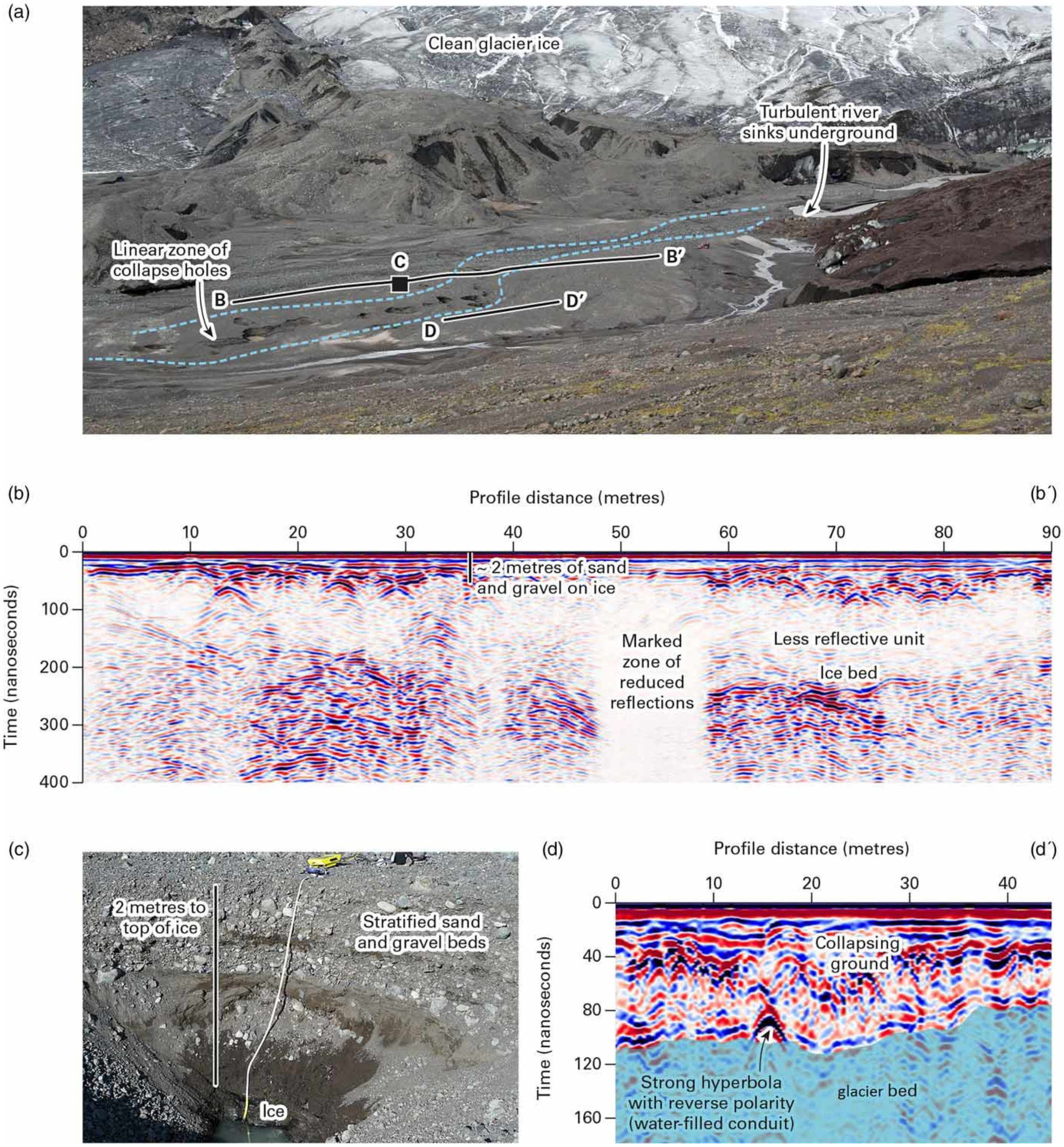

Figure 7 | GPR profiles in the proglacial area. (a). Photograph showing the location of GPR profiles B–B′ and D–D′, and observation Point C. The dashed line represents the hypothetical location on the subglacial buried ice conduit. (b) Un-migrated GPR profile across the proglacial area. The marked zone of reduced reflections coincides with the linear track of collapse features in the photograph in (a). (c) Field photograph showing ~2 m of stratified sand and gravel overlying buried ice. (d) Un-migrated GPR profile showing a collapse structure and fill material, (c) overlying a strong hyperbola interpreted as a water-filled conduit.

systems which have been demonstrated to be open and channelised are often substantially lower than this (e.g., Worthington & Ford 2009), illustrating that it is possible for channelised conduit flow to occur at lower velocities. Also, large diurnal changes in velocity have been measured within the same glacial conduit system and attributed to different flow conditions in the same conduit (Schuler *et al.* 2004). Therefore, changes in velocity and/or dispersivity do not necessarily mean that there has been a change in geometry, and lower

Table 2 | Summary of the findings from selected dye tracer studies through glacier conduits

Velocity (m s^{-1})	Time to peak (min)	Distance (km)	Number of successful tests	Glacier	Region	Publication
0.2–1.5	No number	No number	57	Pasterzengletscher	Austria	Burkimsher (1983)
0.008–0.228	No number	No number	15	Midtalsbreen	S. Norway	Willis *et al.* (1990)
0.085–0.157	75–255	Various	16	Brewster glacier	New Zealand	Willis *et al.* (2009)
0.047–0.32	35–485	0.485–3.335	No number	South Cascade	USA	Fountain (1993)
0.07–0.72	20.35–39.60	3.3	415	Haut Glacier d'Arolla	Switzerland	Nienow *et al.* (1998)
0.04–1.49	No number	1.5–14	43	Leverett glacier	Greenland	Cowton *et al.* (2013)
0.6–1.7	23.7–189	2.1–4.3	12	Gangori glacier	Himalaya	Pottakkal *et al.* (2014)
0.07–0.88	45–240	0.6	9	Rieperbreen	Svalbard	Gulley *et al.* (2012)
0.07–0.58	43–318	1.5	2	Virkísjökull	SE Iceland	This study

velocities and higher dispersivity can occur in glacial systems without the occurrence of distributed flow (Gulley *et al.* 2012). Since it is not possible to classify systems as distributed or channelised from tracer data alone, many studies simply classify glacial systems as having 'fast' or 'slow' flow (Theakstone & Knudsen 1981; Willis *et al.* 2009; Nienow 2011).

In the present study at Virkisjökull, there are several factors that suggest the measured change in velocity need not be attributed to a transition from channelised flow to distributed flow between the two tracer tests. The glacier at the lower ablation zone is thin, and ice movement minimal (Phillips *et al.* 2014), therefore conduit creep closure is slow; in addition, closing by freezing is unlikely as the glacier is temperate and ice melt is recorded all year round in the isotopic composition of the meltwater (MacDonald *et al.* 2016). Data from the proglacial river demonstrate that recession from discrete flood events are the same in both summer and winter (Figure 3, and MacDonald *et al.* 2016), which suggests a perennially active drainage system sufficiently large and channelised to accommodate large flows. The GPR data also suggest that the conduit system persists. These data, from spring 2012 (also at the end of the winter season), suggest that the lower part of the eastern arm of the glacier is characterised by a structurally influenced, interconnected drainage system which is fed by a north–south trending line of moulins on the eastern side of the glacier, which discharges through the major outlet on the eastern side. These observations at Virkisjökull differ from some other studies of temperate glaciers where the arborescent drainage system has been demonstrated to close during the winter (Fountain & Walder 1998).

3. The difference in flow between the two tests may have caused the change in velocity. In this case, the flow could have been through the same conduit system. The flow in the proglacial river was considerably lower in the second test (2 $m^3 s^{-1}$ compared to 5 $m^3 s^{-1}$). Flow in the injection moulin was also substantially lower based on a visual estimation (Figure 2). If there are phreatic sections within the conduit system, then a decrease in flow would result in a decrease in hydraulic gradient causing a decrease in velocity. It is unclear whether there are phreatic sections, although observations suggested that both the injection moulin and the glacier outlet were vadose in the May, August and September visits, suggesting that they remain so throughout the year. Even if the entire system was vadose during the lower flow period of the May tracer test, the lower flow would still result in lower velocities due to increased dispersion and pooling within the channel. Tracer has been shown to move slower and with greater dispersion in channels at low flows as it is slowed down by pools and increased tortuosity around boulders (Hauns *et al.* 2001; Benn & Evans 2014). At high flow, boulders are completely submerged thereby reducing the amount of back-eddy current and temporary storage (Gulley *et al.* 2012). Nienow *et al.* (1996) showed that velocities in glacial conduits were lower during repeat tests in which the flow in the injection moulin was lower, while velocities did not seem to relate to changes in flow in the outlet channel.

This may be because flows in moulins are so much smaller than those in the main conduit system, therefore they are more affected by pooling and debris effects.

4. There could have been a decrease in hydraulic head if there was downwasting of the glacier surface while the outlet remained at the same elevation. This is a possibility given the fast rates of glacier ablation measured annually during repeated visits to the field site (Flett 2016) and the observed stagnation of the lower reaches of the ablation zone. However, a more detailed study of annual and seasonal changes in glacial water table would be required in order to confirm this as a significant effect.

In conclusion, it is not possible to determine with certainty the cause of the substantially lower velocity in the second tracer test from the moulin on the east arm of the glacier, and any of the four possibilities discussed above may be occurring. However, it seems most likely that it is a consequence of the reduced flow in the injection moulin which resulted in more pooling and dispersion, leading to a breakthrough curve with much more tailing, and a lower velocity, but one which is still comparable with velocities observed in groundwater karst conduit systems.

The proglacial foreland drainage system

Both the tracer test and GPR indicate that the buried ice appears to have retained its main meltwater conduit enabling rapid transport of glacial meltwater through the proglacial area. GPR data from the proglacial area demonstrate the presence of an extensive mass of ice buried beneath 1–2 m of outwash sands and gravels. Cavities and conduits are evident within the buried ice, and a significant drainage route is also indicated on the surface by the presence of collapse features.

The tracer test through this area also suggests that drainage is via a conduit system. Water emerging from the Virkisjökull glacier terminus rapidly sinks into the buried ice in the foreland via a large kettle hole and flows through a conduit system in the buried ice to re-emerge within the proglacial lake. The presence of dye tracer specifically on the western side of the lake outlet channel, and not on the eastern side, suggests that discharge occurred at a localised point rather than in a dispersed manner, and that once meltwater emerged from the conduit system into the lake, dispersion within the lake was minimal. If there was a substantial reduction in flow within the lake the tracer would have become too dispersed and diluted through the lake area to be able to detect it at the outlet. This was supported by visual observations of a fast flowing channel that was visible, within the lake, and on the east side which seemed to supply the east side of the outlet channel, where the fluorometer was stationed (Point 5 on Figure 1). It seems likely that at the time of the tracer test there was a conduit within the buried ice which discharged beneath the lake. Although it is unclear how permanent this drainage configuration is, its location in a stagnant area of buried ice, the continuous yearly meltwater supply and GPR profiles suggest that it could be a feature that is exploited for melt water flow throughout the year.

Tracer injected into moulins on the western side of the glacier was not detected in the glacier terminus outlet or at the lake detection points. The flow of water observed in this terminus outlet stream (Point 3 on Figure 1) is substantially less than the flow of water from the lake (Point 4/5 on Figure 1). This suggests that the drainage from the western arm of the glacier may be connected to the proglacial area through a different route. It is possible that this meltwater discharged in a dispersed manner into the lake, diluting the dye to below detection at the lake outlet. However, the injection quantity was much smaller than that in the successful proglacial test which resulted in relatively low tracer concentrations, so even if tracer was discharged into the lake at a specific point, it could have been diluted to below the detection threshold.

The GPR and dye tracing results from the proglacial lake suggest that there is a conduit within buried ice in the proglacial area. This conduit may be the remains of the original subglacial conduit that has been buried in the foreland after terminus retreat. A simplified conceptual model of the evolution of the proglacial area is presented in Figure 8. Ice is buried by the accumulation of debris transported from higher reaches (where ice is still flowing as a result of the steep gradient of the ice fall) (Figure 8(1)). The remains of active meltwater channels, that exploit planes of weakness in the ice, begin to collapse back due to being covered by only a thin layer of ice and sediment. This exposes the

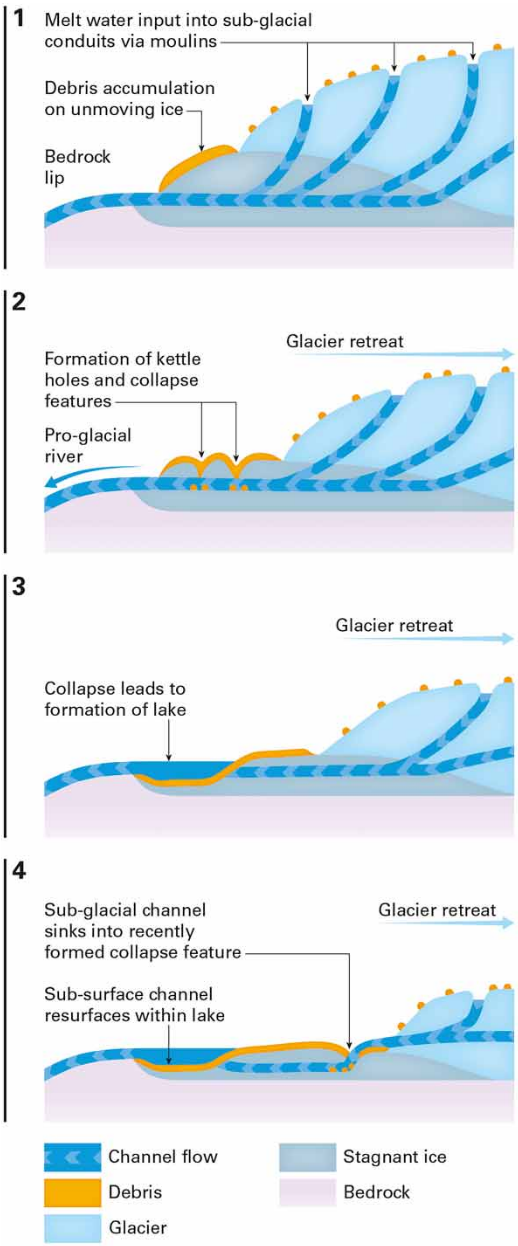

Figure 8 | Diagrams showing conceptually how rapidly retreating glaciers produce a transitional environment and how they evolve and change. In the first stage of deglaciation the slowing of the glacier ice results in unmoving (stagnant) ice at the terminus of the glacier that is subsequently buried by the accumulation of debris transported by the active glacier margin. Meltwater is input into this system from conduits that remain active in the stagnant ice (1). The remains of these active conduits within the buried ice will then begin to collapse to expose water moving through the proglacial buried-ice area (2). This process allows the formation of a proglacial lake that sits upon ice as the active glacier margin continues to retreat (3). The unmoving buried ice is insulated from rapid melting by the accumulation of debris. In the final stage (currently observed at Virkisjökull), the collapse of the active ice margin has exposed an englacial conduit. The meltwater, rather than flowing across the surface, exploits a collapse feature within the foreland to sink back into the conduit system within the buried ice to resurface within the newly formed proglacial lake system (4).

water moving through the area (Figure 8(2)) and a lake begins to form where ponding of water and the formation of surface pools occur (Figure 8(3)). The current proglacial region has a surface river that sinks back below buried ice into a conduit that was formerly connected to the active glacier system. This conduit connects to the lake that formed as a result of the collapse and decay of the ice on the far side of the proglacial area (Figure 8(4)).

CONCLUSION

Tracer testing in the glacial and proglacial areas of Virkisjökull indicates velocities of meltwater near the end of the melt season (September) of $0.58\,\mathrm{m\,s^{-1}}$ and $0.03\,\mathrm{m\,s^{-1}}$, respectively. Meltwater velocities through the glacier are similar to those in the river demonstrating that the conduit system is a highly efficient means of transporting water. The tracer tests suggest the presence of well-developed conduits and channelised flow through both the subglacial and proglacial area. A repeat tracer test in the main glacial system at the end of the winter (May) demonstrated a slower velocity of $0.07\,\mathrm{m\,s^{-1}}$. However, the hydrological behaviour of the river and GPR results suggest that the conduit systems in the glacial and proglacial areas may remain open and active throughout the year.

Rapidly deglaciating catchments such as Virkisjökull create a proglacial setting which is transitional, resulting in an extensive system of buried ice containing the relic conduits of the former ablation zone through which meltwater can be transferred rapidly to the river. Tracer testing and GPR at Virkisjökull have shown that despite the presence of a large lake, meltwater is rapidly transported through the proglacial area to the river. Buried ice in proglacial forelands is likely to become more common as a result of deglaciation, and understanding the hydrology of these areas is important to enable appropriate catchment modelling and hazard mitigation.

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